Dark Material in the Polar Layered Deposits on Mars

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Abstract

Viking infrared thermal mapping (IRTM) and bistatic radar data suggest that the bulk density of the erg material is much lower than that of the average Martian surface or of dark dunes at lower latitudes. We have derived a thermal inertia of 245-280 $Jm^{-2}s^{-½}K^{-1}$ (5.9-6.6 × 10^{-3} cal $cm^{-2}s^{-½}K^{-1}$) for the Proctor dune field and 25-150 (0.6-3.6) for the north polar erg. These data are consistent with the dark material being composed of filamentary sublimate residue (FSR) particles derived from erosion of the layered deposits. The uniqueness of the thermophysical properties of the north polar erg material may be due to a unique polar process that has created them. The visible and near-infrared spectral reflectance of the erg suggests that the dark material may be composed of low-density aggregates of basalt or ferrous clays. Dark dust may be preferentially concentrated at the surface of the layered deposits by the formation of FSR particles upon sublimation of water ice. Further weathering and erosion of these areas of exposed layered deposits may form the dark, saltating material that is found in both polar regions. The dark FSR particles could saltate for great distances before eventually breaking down into dust grains, re-mixing with the global dust reservoir, and being recycled into the polar layered deposits via atmospheric suspension.

I. INTRODUCTION

The Martian polar regions contain key information regarding the current and past climate of Mars. It is widely believed that the Martian polar layered deposits record climate variations over at least the last 10 to 100 million years (Murray *et al.*, 1972; Cutts *et al.*, 1976, 1979; Squyres, 1979; Toon *et al.*, 1980; Carr, 1982; Howard *et al.*, 1982b; Plaut *et al.*, 1988), but the details of the processes involved and their relative roles in layer formation and evolution remain obscure (Thomas *et al.*, 1992). Variations in axial obliquity and orbital eccentricity are thought to influence the climates of both Earth and Mars, but are of greater amplitude in the Martian case (Ward, 1974, 1979; Bills, 1990; Touma and Wisdom, 1993). The Earth's hydrosphere and biosphere do not have a current counterpart on Mars, so it should be simpler to determine the causes and history of climate changes on Mars. Knowledge of the geology of the Martian polar deposits is essential in deducing the processes responsible for their formation and erosion, and the mechanisms by which climatic variations are preserved.

The north polar layered deposits on Mars appear to be the source of the dark material that comprises the north polar erg (Thomas and Weitz, 1989). The south polar layered deposits are also probably the source of some of the dark, saltating material in the south polar region. The physical properties and chemical composition of the erg material therefore have important implications for the origin and evolution of the Martian layered deposits. The focus of this study is the recent geologic evolution of the polar layered deposits as inferred from their geomorphology and the physical properties of materials derived from erosion of the deposits. A review of previous work on the layered deposits is followed below by a description of the observations and models used in this study. Interpretations and hypotheses for layered deposit evolution and polar dune composition are then discussed.

II. GEOLOGIC OVERVIEW AND PREVIOUS OBSERVATIONS

Composition

A common presumption among Mars researchers is that the layered deposits are the result of variations in the proportions of dust and water ice deposited over many climate cycles (Cutts *et al.*, 1979; Squyres, 1979), but their composition is poorly constrained in the north (Malin, 1986) and essentially unknown in the south. Malin (1986) estimated that the density of the northern deposits is 1 g cm^{-3}, suggesting that they are mostly ices. However, the topography and therefore volume of the north polar layered deposits is poorly known, and even the optimistic uncertainty of 50% stated by Malin (1986) allows for up to 50% dust. The lack of observed flow features in the layered deposits indicates that glacial flow has not recently occurred, and led Hofstadter and Murray (1990) to conclude that the layered deposits are less than 40% water ice by volume. However, their modeling suggests that nearly pure water ice would not flow either.

Calculations of the sublimation rate of water ice in the polar regions of Mars (Toon *et al.*, 1980; Hofstadter and Murray, 1990; Paige, 1992) indicate that interstitial ice is not currently stable at the surface of the layered deposits because of their relatively low albedo. The present water ice sublimation rate is high enough to erode the entire thickness of the deposits in about a million years. This result suggests that sublimation of water ice from the layered deposits results in concentration of non-volatile material at the surface of the deposits (Hofstadter and Murray, 1990). Such a lag deposit would thermally insulate underlying water ice from further sublimation, stabilizing the layered deposits against rapid erosion. The existence of a stable, competent layer is indicated by slopes of up to 20° in exposures of the south polar layered deposits (Herkenhoff and Murray, 1990b).

The color and albedo of the layered deposits imply that bright, red dust is the major non-volatile component of the deposits. However, the differences in albedo and color between mantling dust and exposures of layered deposits, and the association of dark saltating material indicates that there is at least a minor component of dark material in the deposits (Thomas and Weitz, 1989; Herkenhoff and Murray, 1990a). The albedo of the layered deposits does not necessarily indicate that an insulating dust layer is present, as the observed albedo only constrains the fraction of dust at the surface to be greater than 0.1% by mass if mixed with water ice grains that have radii of 0.1 mm or larger (Kieffer, 1990). The existence of a lag deposit at least a few millimeters thick is more strongly supported by the low apparent thermal inertia of the surface of the south polar layered deposits (Paige and Keegan, 1994). However, a similar thermal-inertia mapping study of the north polar region indicates that water ice is present near the surface of one area in the north polar layered deposits and is subliming into the atmosphere (Paige *et al.*, 1994). Hence, it appears that while the present erosion rate of the south polar layered deposits is low, the north polar layered deposits (at least in some areas) are currently being eroded by ice sublimation. These inferences have important implications for the present water budget on Mars, and are consistent with estimates of the relative surface ages of the north and south polar layered deposits.

Surface Age

Interpretation of the observed polar stratigraphy in terms of global climate changes is complicated by the significant difference in surface ages between the north and south polar layered terrains inferred from crater statistics. This discrepancy may be due, in part, to the greater areal extent of the southern deposits and the smaller size of the south polar residual cap: most of the north polar layered deposits are covered by the residual cap. These differences and the contrasting residual cap compositions (Farmer *et al.*, 1976; Kieffer, 1979; Paige *et al.*, 1990) indicate that the recent geologic histories of the north and south polar layered deposits are dissimilar.

Using medium-resolution Viking imagery, Plaut *et al.* (1988) found several craters in the southern layered deposits, and showed that the surface of the south polar layered terrain has been accumulating craters for at least the past 120 million years. By modeling crater production and obliteration processes, Plaut *et al.* (1988) used the layered terrain crater size-frequency distribution to obtain an upper limit on a steady-state accumulation of layered terrain materials of 8 km/Ga. This rate is considerably lower than that suggested by Pollack *et al.* (1979) based on observations of dust storm activity, and implies that the south polar layered deposits have accumulated much more slowly than previously thought (*e.g.*, Cutts, 1973; Tanaka and Scott, 1987).

In contrast, Cutts *et al.* (1976) found no fresh impact craters larger than about 300 meters in summertime images of the north polar layered deposits. More recent work using springtime images (in which uniform frost cover and clear atmospheric conditions make craters much easier to recognize) has confirmed the lack of impact craters larger than 100 meters over most of the north polar layered deposits (Herkenhoff *et al.*, 1997). Hence, erosional and/or depositional processes have been more active recently in the north polar region than in the south. The greater extent of aeolian erosional features in the south polar layered terrain may be evidence that gradual erosive processes (such as aeolian abrasion) have been more important than rapid ice sublimation in the evolution of the south polar layered deposits. Furthermore, the inferred average surface age of the south polar layered deposits (at least 10^8 years) is much longer than the timescales of the theoretical orbital/axial variations (10^5 to 10^6 years) that are thought to have induced global climate changes on Mars (Kieffer and Zent, 1992). At least some areas of the south polar layered terrain have therefore not been greatly modified by global climate changes over the last 100 million years or so.

Photogeologic Overview

Layering has been exposed by erosion of gently sloping trough walls in both polar regions (Howard *et al.*, 1982b; Herkenhoff and Murray, 1990b). Howard *et al.* (1982b) also found steep, arcuate scarps eroded into the north polar layered deposits. These scarps are a few hundred meters high and appear to be sources of the dark material that forms dunes nearby (Thomas and Weitz, 1989). The dark material must be a minor component of the layered deposits to explain the significant difference in color and albedo between the dark dunes and the bright red layered deposits (Herkenhoff and Murray, 1990a). If the dark material is sand size, it must have been transported into the polar regions by saltation, and the atmospheric circulation must have been different than that inferred from extant wind indicators (Thomas, 1982). Although the south polar layered deposits may be the source of dark, saltating material in the southern hemisphere, the recognition of source regions is hindered by the lack of high-resolution images (Thomas and Weitz, 1989). Steep scarps have been found in the southern layered deposits (Herkenhoff, 1998; Figure 1), but they do not appear to currently be the source of dark saltating material as in the north polar region.

Herkenhoff and Murray (1990a) identified and mapped five surface units in the vicinity of the south polar residual cap based upon their color (red/violet ratio) and albedo, as deduced from a color mosaic of Viking Orbiter 2 images. A similar technique has been used in conjunction with high-resolution Mariner 9 images to map the geology of three 1:500,000 scale quadrangles: MTM -90000 (Herkenhoff and Murray, 1992), MTM -85080 (Herkenhoff and Murray, 1994), and MTM -85280 (Herkenhoff, 1998). Much of the layered deposits and surrounding terrains appear to be mantled by bright dust, presumably deposited from atmospheric suspension. The interpretation of bright red areas as being covered by dust is supported by their low apparent thermal inertias (Paige and Keegan, 1994) and by the lack of color or albedo contrasts at the boundary of the south polar layered deposits (Herkenhoff, 1998). The dust mantle appears to have been removed from some areas, exposing the darker, less red layered deposits. The distribution of dust in the areas mapped is variable on timescales of weeks to years, as indicated by changes in the extent of the color/albedo units seen in Mariner 9 and Viking Orbiter images. The presence of dark "streaks" in MTM -85280 is consistent with the dark material being mobile as well (Herkenhoff, 1998; Figure 2).

Some areas adjacent to the residual frost cap are brighter and less red than the layered deposits, indicating that frost is present below the limit of resolution of the images. Comparison of Mariner 9 imaging and IRIS data suggests that CO_2 frost was present in these areas, cooling the surface below its radiative equilibrium temperature (Paige *et al.*, 1990). In some areas, the presence of frost late into the southern summer appears to stabilize dust deposits, resulting in net deposition (Howard *et al.*, 1982b, Herkenhoff and Murray, 1990a).

By combining stereophotogrammetry and photoclinometry, Herkenhoff and Murray (1990b) showed that the layered appearance of an exposure of the south polar deposits is due mainly to "staircase" topography rather than albedo variations. It was possible to determine both topographic and albedo variations along photoclinometric profiles using two images of the same area that were taken with different illumination directions. Blasius *et al.* (1982) presented evidence for both topographic and albedo variations between layers in the north polar layered deposits, based on photoclinometric analysis of springtime images (Howard *et al.*, 1982a).

A 17-km impact crater in the south polar layered deposits appears to have been partly eroded since its formation (Figure 3). The preservation of secondary craters and the southern crater wall indicate that little or no "glacial" flow has occurred since the impact (Herkenhoff, 1995). The complete lack of secondaries and crater wall on the north side of the structure indicates that erosion in this area has progressed mainly laterally (north to south) rather than vertically. This is consistent with the hypothesis that erosion of the layered deposits occurs fastest on equator-facing slopes (Howard *et al.*, 1982b; Squyres, 1979).

Dark Saltating Material

Numerous studies have sought to understand sand transport and dunes (accumulations of particles transported by saltation) on Mars from a theoretical standpoint and from Mariner and Viking Orbiter images and thermal data (*e.g.*, Cutts and Smith, 1973; Edgett and Christensen, 1991). Dunes on Mars occur primarily in two geographic regions. The northern circumpolar erg (sand sea) appears as a dark crescent around the north polar residual ice cap. It is the largest concentration of dunes on Mars, with a total area of 7×10^5 km^2 (Tsoar *et al.*, 1979). Transverse dunes cover approximately 50% of the area with an estimated equivalent sediment thickness of 3-4 m (Lancaster and Greeley, 1990), and barchan dunes occur at the margins of the erg. Observed wind streaks and the results of Mars Global Circulation Model simulations suggest that the erg is latitudinally trapped by seasonally-reversing winds (Tsoar *et al.*, 1979; Lancaster and Greeley, 1990) and by winds generated by their relatively low albedo (Thomas and Gierasch, 1995). At several locations at high (40°-75°S) southern latitudes there are groups of impact craters containing dark sediment; some of these intracrater features have been identified as dune fields. These deposits appear to be topographically trapped inside impact craters and are as thick as 100 m (Thomas, 1982). The presence of transverse but not longitudinal dune morphologies in the north polar erg implies geological immaturity through analogy with terrestrial dunes (Tsoar *et al.*, 1979). The pristine nature of Martian dunes and their rapid shedding of bright dust after storms suggest that some are currently active, at least seasonally (Tsoar *et al.*, 1979; Thomas and Weitz, 1989).

Little is known about the primary sources or physical properties of Martian intracrater dune fields, but their apparent thermal inertias have been interpreted in terms of solid grain sizes (Edgett and Christensen, 1991). Edgett and Christensen (1994) found that the thermophysical properties of intracrater dunes are similar within regional clusters but vary between clusters. They found some correlations between dune properties and the local sand supply and/or wind regime. The dark dunes in the north polar erg have inferred thermal inertias (Paige *et al.*, 1994) that are significantly lower than those inferred for dune fields elsewhere on Mars (*e.g.*, Edgett and Christensen, 1994). Because these studies used different thermal models, we have used a single model to compare the apparent thermal inertias of north polar and non-polar dune fields, as described in the next section.

III. THERMAL INERTIA MODELING

The primary goal of our modeling is to derive and compare the thermal inertias of southern mid-latitude dunes and the north polar erg. Thermal inertias derived from spacecraft observations provide one basis for comparing dunes properties (Kieffer, 1973; Edgett and Christensen, 1991). The thermal response of a surface to solar forcing is controlled by its albedo, emissivity, and thermal inertia. The last is the composite quantity $I = \sqrt{k\rho c}$ where k, ρ and c are the bulk thermal conductivity, density, and heat capacity of the surface material. Because the mean free path of CO_2 molecules at Martian surface pressures is comparable to the pore space between sand-size particles, bulk thermal conductivity varies with particle size on Mars. The derived thermal inertia of an unconsolidated surface, then, is a proxy of average particle size assuming a certain grain composition. This relation holds especially well for dune fields, where by terrestrial analogy there is a narrow range of particle sizes.

Thermal inertias are derived by fitting observed brightness temperatures to a model. Error is determined by the quality of the dataset and the accuracy and completeness of the model. An ideal dataset consists of brightness temperatures spanning all local times and seasons and an independently measured albedo and IR emissivity. With sparse spacecraft coverage, a single predawn observation may been used with an assumed albedo and emissivity (Kieffer *et al.*, 1973). Recent studies have shown that the Martian atmosphere significantly complicates the remote derivation of thermal inertia by both heating the surface (Haberle and Jakosky, 1991) and modifying the emitted surface energy flux sensed at the top of the atmosphere (Paige *et al.*, 1994). Viking-era surface thermal models either ignored atmospheric effects or reduced them to a constant surface "heating" term. A discussion of the atmospheric effects on the remote derivation of thermal inertias is given in a following section.

Taken at face value, previously derived thermal inertias of southern mid-latitude intracrater dunes and the north polar erg suggest that they are fundamentally different in composition and/or particle size. In this paper we will present thermal inertias in SI units ($Jm^{-2}s^{-1/2}K^{-1}$) with modified cgs units (10^{-3} cal $cm^{-2}s^{-1/2}K^{-1}$) in parentheses. Correcting for atmospheric effects using a constant heating term gives an inertia of 330-355 (7.9-8.2) for the intracrater dunes (Edgett and Christensen, 1991). Including the time-dependent heating of the surface by a dusty atmosphere, Edgett and Christensen (1994) found the mean thermal inertia of the intracrater dunes to be 270±80 (6.5±1.9). Paige *et al.* (1994), not accounting for any atmospheric effects, found an average thermal inertia of 192 (4.6) for the polar erg. They estimated that the true average could be as low as 96 (2.3). Because different techniques and atmospheric correction schemes were used in these studies, the actual difference in thermal inertia between the intracrater and polar dunes is uncertain.

Correction for Atmospheric Effects

The Martian atmosphere convolutes the derivation of surface thermal inertia from measured brightness temperatures in two ways. For a surface with a given thermal inertia, atmospheric radiation causes a *real* decrease in the surface temperature variation, resulting in an over-estimation of the actual thermal inertia. Radiatively active species in the atmosphere including aerosols absorb solar energy during the day and warm the surface at night. In addition, aerosols cause an *apparent* decrease in the surface diurnal temperature variation as measured at the top of the atmosphere, also resulting in an over-estimation of surface thermal inertia. Measured radiances are decreased during the day and increased at night.

Through the Viking era, atmospheric surface warming was parameterized in Mars thermal models as a constant equal to 2% of the local noontime insolation (Neugebauer *et al.*, 1971). Haberle and Jakosky (1991) constructed a one-dimensional radiative-convective model to calculate the downward radiative flux from a dusty, CO_2 atmosphere and the sensible heat exchange between the surface and atmosphere. They found that even when dust-free, atmospheric warming of the surface considerably exceeds the constant 2% assumption. In addition, the downward flux is minimum near sunrise and maximum near sunset, effectively buffering the surface temperature variation. When only this effect is included, a similar model constructed by Paige *et al.* (1994) yields identical results.

Paige *et al.* (1994) found that the upward radiative flux from the atmosphere tends to further decrease the diurnal temperature variation apparent to the Viking IRTM sensors. The one-dimensional radiative-convective model of Paige *et al.* (1994) used in this paper accounts for atmospheric scattering, emission, and heat transfer between the surface and atmosphere. It also calculates brightness temperatures in each of the IRTM bands at the top of the atmosphere including the effects of atmospheric aerosol absorption and emission.

Aerosol optical properties, vertical distribution, latitude, and season are all important parameters when remotely determining thermal inertias. Pollack *et al.* (1979) and Clancy and Lee (1991) derived dust optical properties from Viking Lander 1 and Viking IRTM, respectively. We find that models using the Pollack *et al.* (1979) dust properties produce derived surface albedos that are significantly higher than independently measured values. Using the Clancy and Lee (1991) dust properties lowers model temperatures for a given surface albedo, correcting this problem. We use the Clancy and Lee (1991) single-scattering albedo and asymmetry parameter in this study. We assume that the dust is uniformly mixed with height. Future Mars missions with simultaneous surface and atmospheric measurements will make more accurate determinations of surface thermal inertia possible.

Methods and Results

Here we present new estimates of the thermal inertias of the intracrater and polar dunes derived using Viking Infrared Thermal Mapper (IRTM) data and the coupled surface-atmosphere model of Paige *et al.* (1994). The dune field within impact crater Proctor (-47°.8S, 330°.3W) and a region within the north polar erg were chosen as study areas. Both sites have been studied previously and appear to be completely covered by dune material (Lancaster and Greeley, 1990; Edgett and Christensen, 1994). IRTM 20 μm observations of these areas were selected using the following general constraints: 1) The IRTM footprint must be smaller than the dune field. 2) The emission angle must be less than 70° to avoid anomalous radiative effects. 3) The times of known dust storms and local frost cover (inferred from high IRTM visible albedo) must be avoided.

Intracrater Dune Field:

We limited our study of intracrater dunes to the dune field within Proctor because it appears to completely cover the crater floor beneath it and has been shown to be representative of intracrater dune fields (Edgett and Christensen, 1994). Shown in Figure 4, it presently serves as a point of comparison with the polar erg. Because the intracrater dune fields are features comparable in size to only the highest resolution IRTM data, very few observations were found that met our constraints. There is little local time coverage in the available Proctor data so we derive thermal inertias using three predawn observations. These IRTM data, listed in Table 2, are from a single Viking Orbiter 1 sequence at L_s=21°.1 and have appeared in other studies (Edgett and Christensen 1991; 1994). The emission angle is 30°, so that each footprint is an ellipse with major and minor axes of 27.5 and 23.7 km respectively.

The surface-atmosphere model was run using a range of input surface thermal inertias and atmospheric opacities. Each run used a visible albedo of 0.12, an infrared emissivity of unity, a surface pressure of 7 mb, and the location, local time, and season of the IRTM observations. The albedo is a typical IRTM solar channel value for daytime observations of the same location. The model outputs 20μm brightness temperatures at the top of the atmosphere, accounting for the emission angle of the observations. Figure 5 shows model 20μm brightness temperatures plotted as a function of thermal inertia and atmospheric opacity. For any opacity between 0.1 and 0.7, the derived thermal inertia lies between 245-280 (5.9-6.7). The near-zero slope of the contours implies that brightness temperature is mostly determined by surface thermal inertia for this specific case. Martin and Richardson (1993) derived a visible opacity of 0.55 from IRTM data at a similar season and location. Table 3 compares our results with those of previous studies for the dune field within Proctor. Our derived thermal inertias are not the lowest, as might be expected, because we use the dust properties of Clancy and Lee (1991).

North Polar Erg:

The north polar erg study site is bound by 80°N-82°N latitude and 135°W-210°W longitude. A Viking Orbiter image taken within this area is shown in Figure 6. Seasonal frost covers this area during the short periods each year when the sun rises and sets daily. The presence of continuous sunlight when the dunes are exposed precludes deriving their thermal inertia using a single predawn temperature observation. Instead, a set of measured brightness temperatures with sufficient local time coverage to define the diurnal temperature variation is fit to model-calculated curves. As noted in Paige *et al.* (1994), the north polar region received excellent spatial and diurnal coverage by Viking during northern summer. We found 1278 IRTM observations between L_s = 98° and L_s = 121° that met our general constraints. The IRTM footprints ranged between 21 to 34 km. These observations are the highest-resolution, lowest-emission angle data of this region. We chose to divide the IRTM data into four 0.5° latitude bins in order to test the uniqueness of the results. Each bin contains approximately 300 observations at various local times, dates, and longitudes. The large range of longitudes (135°-210°) within each bin is necessary to preserve local time coverage.

The large number of observations allowed us to use a fitting procedure similar to Paige *et al.* (1994). The goal is to find the model thermal inertia and albedo that produce the best fit to an entire bin of IRTM data. During the fit, each data point is compared with the model-calculated brightness temperature at the latitude, local time, and season of the observation. The appropriate model brightness temperature is found by linearly interpolating between model outputs at discrete intervals of latitude, local time, and season. The surface-atmosphere model of Paige *et al.* (1994) was run for all combinations of latitude between 80°N and 82°N, thermal inertia between 25 and 275, and albedo between 0.1 and 0.4 with resolutions of 0.5°, 50, and 0.1, respectively. The models were output every hour on Julian dates 2443670, 2443685, 2443700, and 2443715, corresponding to $L_s = 98°$, 105°, 112°, and 119°. All model runs assumed an emissivity of unity, a surface pressure of 7 mb, and an emission angle of 30°. The fitting procedure was repeated for each latitude bin and for model visible atmospheric opacities of 0.0, 0.2, 0.4, and 0.6.

The fitting procedure calculates the standard deviation (sigma) between the IRTM data and the model brightness temperatures over the parameter space of thermal inertia and albedo. The best fit (smallest sigma) for each opacity for one of the latitude bins is given in Table 4. The four latitude bins give similar solutions, increasing our confidence in the uniqueness of the fit. A contour plot showing how sigma varies over the parameter space is presented in Figure 7. In general, there is more uncertainty in the best-fit inertia than albedo. We can use the discrepancy between the derived albedo and the measured IRTM solar channel albedo to estimate the true atmospheric opacity during the IRTM observations. The measured albedo is 0.1-0.2 (Paige *et al.*, 1994), nearest the derived albedo when the model opacity is 0.2-0.4. The derived albedos when the model opacity is 0.0 or 0.6 are significantly different than the measured values. Following this reasoning, our best-fit thermal inertia for the north polar erg is 75 (1.8), with slightly less good fits achieved for all inertias between 25 (0.6) and 150 (3.6).

Error Sources

Our results are robust to several sources of error. Sensitivity tests show that variations of model albedo (±0.1), emissivity, and elevation do not diminish our conclusions. Christensen (1982) found a correlation between low-albedo materials (like dune-forming materials) and lower emissivities. Changing the model emissivity to 0.9 slightly narrows the difference between the intracrater and polar dunes. We assumed that both sites are at the 7 mb pressure level. Smith and

Zuber (1996) determined that the intracrater dunes lie a few kilometers above this level, while the polar erg lies a few kilometers below. Bridges (1994) studied the effect of elevation on derived thermal inertias. He found that lower (higher) elevation surfaces receive more (less) atmospheric warming, resulting in lower (higher) derived inertias when properly modeled. The magnitude of the correction is small and the sign of the correction reinforces our conclusions: the lower-elevation polar erg may have an even lower inertia, while the intracrater dunes may have a higher inertia. This correction is used in the calculations in the next section.

Uncertainties in dust optical properties have significant consequences. The inertias derived for the Proctor dune field are lower by as much as 120 (3) if we assume the dust properties of Pollack *et al.* (1979) instead of Clancy and Lee (1991). With the single-point derivation method, however, systematically high model surface temperatures also will produce low derived inertias. The fact that the derived albedo of the polar erg greatly exceeds measured values when the Pollack *et al.* (1979) properties are used supports the idea that these properties produce systematically high model temperatures.

IV. DISCUSSION AND INTERPRETATIONS

We have derived a thermal inertia of 245-280 (5.9-6.6) for the Proctor dune field and 25-150 (0.6-3.6) for the north polar erg. Both results are lower than most previous studies because we have dealt more fully with atmospheric effects that tend to increase derived inertias. Nevertheless, our conclusion is that the thermal inertias of intracrater dune fields are significantly higher than those of the north polar erg. The difference is greater than the variation among individual intracrater dunes (Edgett and Christensen, 1994). This result implies that the physical properties of the polar dunes are indeed fundamentally different than those of non-polar dunes.

However, the dark material that has been transported to, incorporated in, and eroded from the polar layered deposits, and at least partly form the polar erg, may have the same chemical composition as the globally distributed dune-forming material. This hypothesis is supported by the fact that the polar and intracrater dunes, although separated greatly in latitude, have approximately the same color and albedo (Thomas and Weitz, 1989). Bell *et al.* (1997) interpreted absorption features near 953 nm in Hubble WF/PC2 images of the north polar erg as evidence for pyroxene in the coarse, basaltic particles that form the dunes. If the dark material is composed of solid sand-size grains (hereafter called "sand" without compositional implications), poleward circulation is required to transport the sand (by saltation) into the layered deposits (Thomas, 1982). Saltating sand would eject dust into suspension, hindering co-deposition of sand and dust. Aeolian clay deposits in Australia that contain solid mineral grains have been described (and called "parna") by Butler (1956), but the clay fraction of these deposits is emplaced by saltation of "stable silt or sand-sized clay aggregates" (Dare-Edwards, 1984). A similar mechanism may have allowed dark sand grains to be emplaced with dust aggregates by saltation into the polar layered deposits. Alternatively, sand may have saltated over ice-cemented dust toward the poles at some previous time when winds blew onto the polar caps. In this case, the dark sand must have formed layers or lenses less than a few meters in size, or they would be visible in high-resolution Viking Orbiter images.

In any case, transportation of sand into the polar layered deposits by saltation from distant sources is problematic, because sand grains are not expected to survive saltation over large distances due to the high speed of the winds required to move them (Sagan *et al.*, 1977; Krinsley and Leach, 1981; Greeley *et al.*, 1982). In laboratory experiments with fine augite and basalt grains, Krinsley and Leach (1979) found that particles are rapidly broken into silt and clay-size particles under expected Martian wind conditions (velocities greater than 20 m/sec). This finding supported the calculations by Sagan *et al.* (1977) that predicted the preferential depletion of 150-micron grains due to collisional disruption. More recent single-particle experiments confirm that grains of all sizes are shattered upon impact of a solid target at 100 m/sec (Marshall *et al.*, 1996).

Porous sand-size grains should saltate at lower wind velocities than solid grains (Iversen *et al.*, 1976), and will therefore survive longer before they are comminuted. Basaltic pumice fragments and other pyroclastic debris are likely to exist on Mars, and may form dunes upon aeolian sorting. Such material may be episodically injected into the Martian atmosphere and distributed widely, perhaps into the polar regions. Hence, the dark material in the polar layered deposits may have been deposited during episodes of volcanic activity, and has been more recently exhumed and redistributed. Because of their high surface area/volume ratio, these particles may weather more rapidly than solid grains on Mars, explaining their relative lack in older, low-latitude dune fields.

Alternatively, dark dust (rather than sand) may be intimately mixed with bright dust in the layered deposits, as suggested by Herkenhoff and Murray (1990a). Dark dust in the layered deposits may have formed the dunes observed in the polar regions by aggregation during erosion of the layered deposits. Sublimation of phyllosilicate dust/ice mixtures has been shown to result in the formation of filamentary sublimate residue (FSR) particles of various sizes, even when mixed with grains that do not form FSR (Storrs *et al.*, 1988; Thiel *et al.*, 1991). The dust particles in this FSR appear to be held together by ionic bonds rather than by a cementing material (Storrs *et al.*, 1988) or electrostatic forces as in the aggregates formed in the absence of water by Greeley

(1979) and Krinsley and Leach (1981). Such particles can saltate along the Martian surface, and may therefore create dunes (Saunders *et al.*, 1985; Saunders and Blewett, 1987). However, it seems unlikely that such particles could survive saltation over very long distances into areas far outside the polar regions. The color of the parts of the north polar layered deposits, including the steep scarps that appear to be the source of the dark dunes, is similar to dust deposits elsewhere on Mars (Thomas and Weitz, 1989). Other areas are darker and less red, perhaps due to a lag of dark FSR at the surface or the presence of dark dunes below the resolution of the images. Recent dust mantles may be covering much of the north polar layered deposits, as observed in the south (Herkenhoff, 1998), masking the true color of the layered deposits.

Magnetic dust grains would be expected to form FSR more easily than non-magnetic dust. Experimental formation of FSR with magnetic material has not been attempted, and should be the subject of future research. There is direct evidence for 1-7% magnetic material (magnetite or maghemite) in the surface fines at the Viking lander sites (Hargraves *et al.*, 1979), and titanomagnetite is found in many SNC meteorites (McSween, 1985). In addition, analysis of Viking lander sky brightness data indicates that suspended dust over the landing sites contains about 1% opaque phase, perhaps of the same composition as the magnetic material on the surface (Hargraves *et al.*, 1979; Pollack *et al.*, 1979). Results of the magnetic properties experiment on the Mars Pathfinder lander indicate that the atmospheric dust particles are composite, containing 6% (on average) maghemite or titanomaghemite (Hviid *et al.*, 1997). Pedersen *et al.* (1998) concluded that single-phase particles of maghemite or magnetite are not present as free particles in the airborne dust in any appreciable amount. Within the uncertainties in these measurements, the percentages of magnetic material given above are identical to the volume of dark dune deposits in the polar regions expressed as a percentage of the estimated volume of eroded layered deposits (Table 1). This comparison indicates that the presence of magnetic dust in the layered deposits is likely, and that formation of dunes from magnetic aggregate particles is plausible. Eventual destruction of the saltating particles could allow recycling of the dust into the layered deposits via atmospheric suspension in dust storms. A significant problem with this scenario is that the polar dunes are much darker than the magnetic dust at the Viking and Pathfinder landing sites. While magnetic dust in the polar regions may differ from the dust observed at the landing sites, the difference in albedos leads us to consider alternative hypotheses. However, under the assumption that FSR can be formed by sublimation of mixtures of water ice and magnetic dust, the properties of this material, as well as sand grains, have been estimated and compared with observational data, as detailed below.

Inferred Particle Properties

Previous interpretations of Martian thermal inertia data in terms of particle sizes have utilized the relationship between these quantities presented by Kieffer *et al.* (1973), which is based primarily upon measurements of the thermal properties of quartz sands (Wechsler and Glaser, 1965). The low albedos of Martian dunes are inconsistent with a siliceous composition, so basalt and magnetite particles are considered here (the thermal properties of maghemite have not been reported in the literature). The thermal conductivity k of the observed dunes can be derived from their thermal inertia I using the following equation:

$$k = \frac{I^2}{\rho_0 (1-f) c}, \qquad (1)$$

where ρ_0 is the density of solid basalt (2750 kg m^{-3}) or magnetite (5200 kg m^{-3}), f is the fraction of pore space (porosity) in the dunes, and c is the specific heat of basalt (586 J kg^{-1} K^{-1}) or magnetite (544 J kg^{-1} K^{-1}). This equation was used to generate the values in Tables 5 and 6.

Atmospheric pressure affects the interpretation of thermal inertias (Bridges, 1994), so we have estimated the atmospheric pressure at the seasons and elevations of the two study areas. The seasonal variation in pressure observed at the Viking Lander 1 site (Zurek *et al.*, 1992) was used to calculate the pressure at each of the study areas at the times that they were observed by the IRTM. Based on elevations of the VL1 site and the two study areas shown on the USGS topographic map of Mars and assuming an atmospheric scale height of 10.8 km, the pressure at the north polar study area was about 7 mb (5.3 torr), while the pressure at the Proctor dune field was 4.3 mb (3.3 torr). These pressures are used in the following evaluation of the hypotheses described above.

Solid Grains

Presley and Christensen (1997b) found the thermal conductivity of a basaltic dune sand sample with a grain size range of 60-250 μm (49% porosity) to be the same (within measurement uncertainty) as that of 160-180 μm glass beads (36% porosity) over a range of pressures including those expected at the surface of Mars. Their results showed that mixtures of different particle sizes have the same thermal conductivity as a pure sample of the largest particles, and tentatively concluded that the largest particles in a mixture dominate the thermal conduction across the sample. The thermal conductivity of 44-104 μm (54% porosity) basaltic sands, interpolated at 5 torr from Wechsler and Glaser (1965), is 1.8×10^{-2} W m^{-1} K^{-1}, ~50% less than that of glass beads

90-100 μm in size (36% porosity) and 38% less than 70-75 μm glass beads (Presley and Christensen, 1997a). These differences are primarily due to the greater roughness and lower bulk density of the basalt particles (Presley and Christensen, 1997b). The rougher particles are expected in nature, so basaltic dune particles should have thermal conductivities that are less than those of samples of glass beads of equivalent size.

The thermal conductivity of solid magnetite (Touloukian *et al.*, 1970) is identical (within measurement uncertainties) to that of pure quartzite (Carte, 1955), so the thermal conductivity of magnetite sand is assumed to be the same as that of glass beads or quartz sand. The thermal conductivities of the materials considered here are only weakly dependent on temperature between 200 and 300 K, so values measured near 300 K have been used in all cases.

Proctor dunes:

The thermal inertia of the Proctor dune field indicates that the thermal conductivity of the dunes is 0.06 W m^{-1} K^{-1} for basalt and 0.036 W m^{-1} K^{-1} for magnetite (Table 5, 30-35% porosity). Using the results of Presley and Christensen (1997a) at 3 torr, these values are consistent with basalt grains about 600 μm in diameter, or magnetite grains about 180 μm in diameter. Glass spheres were used in the experiments of Presley and Christensen (1997a), but particle composition does not significantly affect the thermal conductivity of particulate samples (Presley and Christensen, 1997b). Because the Proctor sand particles are probably rougher than glass beads, they are probably somewhat larger on average than the above estimates. These large grain sizes are puzzling, as the wind speeds required to move these particles should accelerate the particles to velocities sufficient to destroy them upon surface impact (Sagan *et al.*, 1977; Krinsley and Leach, 1981). Although Marshall *et al.* (1998) have shown that grains projected into beds of similar sized grains at velocities expected on Mars can survive the impact, the grains must have been collected into dunes by saltation across rocky surfaces. This implies that the coarse sand in the intracrater dunes must have been derived locally. Alternatively, rocky areas between dunes may be contributing to the high apparent thermal inertia of the Proctor dune field. Such areas are not visible in Fig. 4, perhaps because they have the same low albedo as the dunes. High-resolution images and thermal data from the MGS spacecraft will be useful in resolving this paradox.

North Polar Dunes:

The thermal inertia of the north polar dunes indicates that their thermal conductivity is very low, even if they are quite porous. For the least dense glass bead and crushed quartz samples (70-75% porosity) measured by Presley and Christensen (1997a), the polar dune thermal inertia implies thermal conductivities of 0.01 W m^{-1} K^{-1} for basalt and 0.008 W m^{-1} K^{-1} for magnetite (Table 6). Using the results of Presley and Christensen (1997a) at 5 torr, these values are consistent with basalt grains about 8 μm in diameter, or magnetite grains about 4 μm in diameter. While rough particles could be somewhat larger, particles in this size range are expected to be transported by atmospheric suspension (Edgett and Christensen, 1981; Greeley *et al.*, 1994), and are therefore not likely to form dunes. More likely, ground ice is present below the diurnal skin depth with even lower-inertia material at the surface (Paige and Kieffer, 1987). Very porous grains such as basaltic pumice or ash are possible constituents of the north polar dunes, but the thermal conductivity of 44-104 μm pumice grains at Martian pressures reported by Wechsler and Glaser (1965) is very similar to basalt grains of the same size, despite its lower bulk density. Hence, other low-inertia materials that are capable of saltation must be examined as possible dune-forming materials on Mars. The thermal properties of FSR particles are therefore estimated below.

Dust Aggregates

The porosity of clay FSR formed in laboratory experiments is 99% (Storrs et al., 1988). Magnetite FSR would therefore have a bulk density of only 52 kg m^{-3}. The thermal conductivity of porous clay at 313 K, 740 torr ranges from 0.477 to 2.05 W m^{-1} K^{-1}, depending on water content (Touloukian et al., 1970). The lowest value is identical to that of clay FSR (Storrs et al., 1988). When this dry clay was placed in a high vacuum, its thermal conductivity decreased only 7%. Therefore, the conductivity of clay FSR at 6 mbar is probably about 0.47 W m^{-1} K^{-1}. The thermal conductivity of clay *minerals* is probably similar to that of most silicates or magnetite. Hence, magnetite FSR should have a thermal conductivity of 0.5 W m^{-1} K^{-1} or less, implying a thermal inertia of no more than 119 J m^{-2} $sec^{-1/2}$ K^{-1}. The thermal inertia of basalt FSR with the same porosity would be 90 J m^{-2} $sec^{-1/2}$ K^{-1}, also compatible with the north polar erg thermal inertias derived above. While a scenario involving preferential aggregation of magnetic dust grains upon erosion of the polar layered deposits is attractive, the HST observations of a possible pyroxene absorption band in the north polar erg (Bell *et al.*, 1997) are more consistent with aggregates of basaltic dust or ferrous clays (Calvin, 1998). In any case, formation of the north polar erg by dust aggregates is plausible, and can explain the thermal properties of the dunes.

The presence of such aggregates is also consistent with radar observations of the north polar erg. Viking bistatic radar data show that the north polar dunes are rougher and less reflective (and therefore less dense) than surrounding terrains (Simpson and Tyler, 1981).

Increased radar roughness is expected for dune fields, and their low apparent density is consistent with the presence of porous grains or dust aggregates. The only solid materials found by Campbell and Ulrichs (1969) to have permittivities less than 3 are pumice, semi-welded tuff, and volcanic ash. Many types of powdered rocks have permittivities around 2 when their density is 1 g/cm^3. Although magnetite is a conductor, experiments involving low-density suspensions of conductors in dielectrics (Kelley *et al.*, 1953; Pettengill *et al.*, 1988) suggest that the index of refraction (and therefore reflectivity) of magnetite FSR might be very low. However, the Viking radar data are more consistent with low-density, nonconductive particles such as basaltic or ferrous clay FSR.

V. CONCLUSIONS

The thermal inertia of the north polar erg is 25-150 (0.6-3.6), much less than that of the Proctor dune field: 245-280 (5.9-6.6). This difference suggests that the north polar dune material was formed by a different (probably uniquely polar) process than the dune materials at lower latitudes. The low thermal inertia of the north polar dunes implies that they are composed of very low density material, probably aggregates of dust such as the FSR particles formed by sublimation of dust/ice mixtures. Micron-sized basalt or ferrous clay particles are likely components of the aggregates, as they can easily be transported into the polar regions via atmospheric suspension and are consistent with near-infrared spectra of the north polar erg.

The observed thermal inertias of southern mid-latitude dune fields on Mars are consistent with ~180-micron magnetite grains with 35% porosity, or ~600-micron magnetite grains with

30% porosity. The wind velocities required to move particles of this size may be sufficient to destroy them upon impact with a rocky or sandy surface, so either they are locally derived or the apparent thermal inertia is affected by coarser material between dunes.

Weathering of the Martian layered deposits by sublimation of water ice can account for the geologic relationships observed in the Martian polar regions. The non-volatile component of the layered deposits appears to consist mainly of bright red dust, with small amounts of dark dust or sand. Dark dust may preferentially form filamentary residue particles upon weathering of the deposits. Once eroded, these particles may saltate to form the dark dunes found in both polar regions.

New orbital observations of the Martian polar regions from the Mars Global Surveyor and surface exploration by the Mars Volatiles and Climate Surveyor are likely to greatly enhance our understanding of the polar layered deposits and the climate changes that they record. We look forward to the acquisition of data that will clarify the relationships between the layered deposits and the dark dune materials in both polar regions.

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Figure Captions

Figure 1. Part of Mariner 9 image 208B02 of scarp in south polar layered deposits at 83.3°S, 296°W (barely visible near top of image). Directions toward sun and north are indicated. Resolution = 88 m/pixel, no filtering, image is 34 km across. Note that scarp is fluted, with flutes about 1 km across, indicating modification by mass wasting. L_s = 350° (southern summer).

Figure 2. Part of Viking Orbiter image 383B27 of south polar layered deposits, showing dark "streaks" on equator-facing low-relief scarp. Area shown in figure 1 is outlined, which includes scarp that is barely resolved here. Other probable scarps indicated by arrows. Resolution = 190 m/pixel, image is 139 km across. L_s = 328.6° (southern summer).

Figure 3. Viking Orbiter image 421B79 of partly eroded 17-km diameter crater (near right edge) and secondary craters in south polar layered deposits, centered at 85.0°S, 352°W. Orthographic projection with north toward bottom, illumination from bottom, resolution 133 m/pixel. L_s = 348.4° (southern summer).

Figure 4. Mosaic of Mariner 9 narrow-angle images of Proctor crater, at 47.8°S, 329°W. Image is 110 km high, illumination from north (top).

Figure 5. Contour plot of model-calculated 20-μm brightness temperatures versus model atmospheric opacity and thermal inertia for the Proctor dune field. The dashed lines represent the

range of 20-μm brightness temperatures measured by IRTM. The derived thermal inertia is nearly independent of atmospheric opacity.

Figure 6. Viking image 59B32 of north polar dunes, centered at 81.2°, 352°W. Orthographic projection, illumination from bottom right, image 104 km across. $L_s = 134.6°$ (northern summer).

Figure 7. Contour plot of the standard deviation in Kelvins between IRTM and model-calculated 20-μm brightness temperatures versus model thermal inertia and albedo for the north polar erg within 81°-81.5°N and 135°-210°W. The best-fit thermal inertia and albedo are 75 J $m^2 s^{-1/2} K^{-1}$ and 0.16, respectively. The visible atmospheric opacity is assumed to be 0.4.

Table 1.

Quantity	Fraction	Reference
Magnetic material in airborne dust	~6%	Hviid et al., 1997
Magnetic material in surface fines	1-7 %	Hargraves et al., 1979
Opaque phase in atmospheric dust	~1%	Pollack et al., 1979
Volume of dark dunes divided by volume of eroded polar deposits	1-10 % 0.1-1%	Thomas, 1982 Lancaster and Greeley, 1991

Table 2. Proctor Intracrater Dune Field IRTM Data

Lat.	Lon.	Hour	s/c-orb-seq-sp-ick	T_{20} (K)
-47.86	328.45	4.96	1-549-9-762-6	167.99
-47.89	327.77	5.00	1-549-9-763-6	166.61
-47.90	327.20	5.04	1-549-9-764-6	167.73

Table 3. Summary of Thermal Inertias Derived for Proctor Dune Field

Authors	Surface/Atmosphere Model	Visible Opacity	Derived T. I. (mks)
Edgett and Christensen (1991)	Kieffer et al. (1977)	N/A	340
Edgett and Christensen (1994)	Haberle and Jakosky (1991)	0.0 0.4	300 230
Herkenhoff and Vasavada (1998)	Paige et al. (1994)	0.1-0.7	245-280

Table 4. Thermal Inertias for the North Polar Erg, 81°-81.5°N

Visible Opacity	Best-fit Inertia	Sigma (K)	Inertias with Sigma < 2.5 K	Derived Albedo
0.0	75	2.40	50-90	0.32
0.2	75	2.27	25-150	0.24
0.4	75	2.31	25-125	0.16
0.6	25	2.35	25-50	0.10

Table 5. Thermal conductivity of Proctor dunes ($I = 260$)

Porosity	Basalt	Magnetite
25	5.59E-02	3.19E-02
30	5.99E-02	3.41E-02
35	6.45E-02	3.68E-02
40	6.99E-02	3.98E-02
45	7.63E-02	4.34E-02
50	8.39E-02	4.78E-02
55	9.32E-02	5.31E-02
60	1.05E-01	5.97E-02
65	1.20E-01	6.83E-02
70	1.40E-01	7.97E-02
75	1.68E-01	9.56E-02
80	2.10E-01	1.19E-01
85	2.80E-01	1.59E-01
90	4.19E-01	2.39E-01
95	8.39E-01	4.78E-01
99	4.19E+00	2.39E+00

Table 6. Thermal conductivity of north polar dunes ($I = 75$)

Porosity	Basalt	Magnetite
25	4.65E-03	2.65E-03
30	4.99E-03	2.84E-03
35	5.37E-03	3.06E-03
40	5.82E-03	3.31E-03
45	6.35E-03	3.62E-03
50	6.98E-03	3.98E-03
55	7.76E-03	4.42E-03
60	8.73E-03	4.97E-03
65	9.97E-03	5.68E-03
70	1.16E-02	6.63E-03
75	1.40E-02	7.95E-03
80	1.75E-02	9.94E-03
85	2.33E-02	1.33E-02
90	3.49E-02	1.99E-02
95	6.98E-02	3.98E-02
99	3.49E-01	1.99E-01